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A hydrological and water temperature modelling framework to simulate the timing of river freeze-up and ice-cover breakup in large-scale catchments

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ABSTRACT

Ice phenology, defined as the timing of freeze-up and ice-cover breakup, plays a key role in streamflow regimes in cold-region river catchments. River freeze-up and ice-cover breakup events are controlled by meteorological and hydrological variables. In this study, we present a modelling framework consisting of a physically-based semi-distributed hydrological model and the integration of a 1D stream temperature model that can predict the ice duration in cold region rivers. The hydrological model provides streamflow and hydraulic parameters for the stream temperature model to obtain instream water temperature. The model was successfully applied in the Athabasca River basin in western Canada. Calibration was carried out using the water temperature recorded in the stations at the towns of Hinton, Athabasca and Fort McMurray. Model results show consistent correspondence between simulated freeze-up and breakup dates and the hydrometric station data. In the main tributaries of the basin, freeze-up timing spans from the last week of September to the second week of November and ice-cover breakup occurs from the second week of March to the last week of May. The model presents an application of water temperature and ice phenology simulation which can be incorporated in ice-jam flood forecasting and future climate change studies.

1. Introduction

In mid-to high-latitude lakes and rivers, the formation and breakup of ice covers are important seasonal events. River freeze-up and ice-cover breakup events are influenced by stream water and air temperature, and are sensitive to the characteristics of individual waterbodies (Beltaos and Prowse, 2001). In cold region rivers, the ice phenology plays a key role in the effects of ice on river bank and bed erosion, sedimentation and regulation of the stream flow regime in winter (e.g., Beltaos and Prowse, 2009; Terry et al., 2017). Climate influences the timing of freeze onset, ice duration, ice melt and thermal dynamics that feed back to the climate system, initiating further changes.

The ice formation process in rivers is initiated once the air temperature has been below 0 °C long enough to super-cool the water and form frazil ice (e.g., Osterkamp, 1978; Beltaos, 2000). Dynamic, physical and mechanical factors are involved in the formation of ice covers (Turcotte and Morse, 2013). On lentic waters, such as lakes and along river shorelines with low flow velocity, a solid continuous ice sheet forms with a horizontal uniform structure. This ice sheet originates from either the waterbody shores or pre-formed plate ice that extends outward to the center of the waterbody. In rivers with more turbulent flow, the formation is more dynamic and mainly results after the accumulation of frazil slush, pancake ice, and ice floes (e.g., Beltaos, 2000; Prowse and Beltaos, 2002). Accumulation on rivers is either an ice jam or a hummocked ice cover -pressure ice mounds haphazardly arranged-when it refreezes.

The breakup of river ice covers is triggered by mild weather conditions and encompasses a variety of processes associated with thermal deterioration, initial fracture, movement, fragmentation, transport, jamming, and final clearance of ice. During ice-cover breakup, multiple stages such as 'pre-breakup', 'onset', 'drive' and 'wash' occur concurrently within a specific reach (Beltaos, 2008). During the pre-breakup phase, the ice cover becomes more susceptible to fracture and movement via thermal and solar radiation induced reductions in its thickness and strength. At the same time, spring and early summer weather conditions bring about snowmelt and rainfall events that cause an increase in flow discharges and the rise of stream water levels that fracture, drive and wash out the ice. Phenology records obtained over long time periods and at regular time intervals are thus required to better understand inconsistencies and changes in meteorological conditions and how such changes impact river ice processes (Latifovic and Pouliot, 2007). Due to the lack of observed records, stream water temperature models constitute important tools to simulate ice

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phenology data, especially in ungauged catchments.

Several model approaches, such as linear and multivariable regression methods, neural network models and thermal energy budget methods, have been developed and implemented to simulate stream water temperature (e.g., Benyahya et al., 2007; Dugdale et al., 2017). Most of these data-driven models have been applied to specific stream reaches and calibrated using observed data. In contrast, mechanistic models have been applied for regional and catchment scale problems, to understand the spatial and temporal behaviour of the energy transfer mechanisms controlling water temperature.

Stream temperature models have been utilized for short-term (e.g. daily) and long-term (from weeks to years) water temperature forecasting (Dugdale et al., 2017). Similarly, water temperature hindcasting has been performed to reconstruct historical records (Benyahya et al., 2007). Water temperature models are also employed to investigate the potential impact of anthropogenic effects on the thermal regime of rivers (Sun et al., 2015) and the implications for aquatic life and ecosystems (e.g., Gu et al., 1999). Despite the variety of applications, physically-based distributed stream water temperature models have not been used to simulate the timing of freeze-up and breakup events in large river catchments.

In this paper, the River Basin Model (RBM), a semi-Lagrangian water temperature model for advection-dominated river systems (Yearsley, 2012), was coupled to MESH, a semi-distributed hydrological model (Pietroniro et al., 2007), to simulate stream water temperature in the Athabasca River basin (ARB). Although the model has been successfully coupled to the Variable Infiltration Capacity (VIC) model (e.g., Van Vliet et al., 2012), widely implemented in large-scale hydrological modelling, the MESH-RBM system is more suited to cold-region catchments due to its ability to simulate snow fall, snow redistribution and snow melt and their effect on stream water temperature. Due to the lack of stream water temperature data in the upland areas of the ARB, temperature data in these areas was extracted from thermal infrared imagery via remote sensing techniques to impose model boundary conditions at the headwater streams. Simulated stream water temperature time series were used to estimate the timing of river freeze-up and ice-cover breakup events in the ARB. The model may be useful for improving ice-jam flood event preparedness and mitigation, and to study future ice phenology.

2. Methods

2.1. RBM model description

RBM is a mechanistic one dimensional stream temperature model that solves the 1D-head advection equation:

$$\frac{d(T_w A_x)}{dt} = \frac{H_{air-water} w_x}{\rho_w C_p} + \frac{Q_{trb}\Delta T_{trb}}{\delta x} + \frac{Q_{eff}\Delta T_{eff}}{\delta x} \quad (1)$$

where ρ_w is water density (kg m^{-3}); C_p is the specific heat of water (J kg^{-1} ^{o}C^{-1}); $T_w$is water temperature (^{o}C); A_x is the cross section area at distance x (m^{-2}); w_x is stream width at distance x (m); Q_{trb} is the advected flow from tributaries or the subsurface (m^{-3} s^{-1}); ΔT_{trb} is the difference between advected temperature from tributaries or the subsurface, respectively, T_{trb} and T_w(^{o}C); Q_{eff} is the advected flow from head dumps (e.g. thermoelectric power plants) (m^{-3} s^{-1}); ΔT_{eff} is the difference between advected temperature from head dumps (T_{eff}) and T_w (^{o}C); δx is the longitudinal distance along the river axis (m) and t is time (s). The net heat flux at the air-water interface ($H_{air-water}$) is calculated in J m^{-2} s^{-1} as:

$$H_{air-water} = (H_{sw} - H_{rs}) + (H_a - H_{ar}) + H_{evap} + H_{cond} + H_{back} \quad (2)$$

where H_{sw} is short-wave solar radiation (J m^{-2} s^{-1}); H_{rs} is reflected short-wave solar radiation (J m^{-2} s^{-1}); H_a is long-wave atmospheric radiation (J m^{-2} s^{-1}); $H_{ar}$is reflected atmospheric radiation (J m^{-2} s^{-1}); $H_{evap}$is the evaporative heat flux (J m^{-2} s^{-1}); H_{cond} is conductive heat flux (J m^{-2} s^{-1}) and H_{back} is black body radiation from the water surface (J m^{-2} s^{-1}). Although some authors have considered hyporheic flow, riparian vegetation and stream bed head exchange, important factors affecting $H_{air-water}$ (Equation (2)), these are not relevant due to the scale of our study case and the purpose of this research.

RBM uses an explicit semi-Lagrangian numerical scheme (Yearsley, 2009) to resolve Equation (1). The scheme tracks individual water packets along the river network in the flow direction, constituted by interconnected segments and nodes of homogeneous characteristics, and saves the simulated water temperature at every segment of the network.

2.2. Boundary conditions in RBM

RBM requires water temperature boundary conditions at the basin headwaters. Different approaches, such as equalizing water temperature with simulated daily soil temperature from land-surface numerical models and estimating water temperature as a function of air temperature, have been tested before (Van Vliet et al., 2012). The non-linear water temperature regression model described by Mohseni et al. (1998) has yielded acceptable results. This model relates weekly water temperatures as a non-linear function (S-curve) of weekly air temperatures as:

$$T_{w_{head}} = \mu + \frac{\alpha - \mu}{1 + e^{\gamma(\beta - T_{a_{smooth}})}} \quad (3)$$

where μ is the lower bound of water temperature (^{o}C); α is the upper bound of water temperature (^{o}C); $\gamma = \frac{4\tan\theta}{\alpha - \mu}$ is the measure of the slope at the inflection point of the S-shaped relation (^{o}C^{-1}), where tan$\theta$is the slope at the inflection point; β is the air temperature at the inflection point (^{o}C) and T_{whead} is the headwater temperature (^{o}C). As water temperature variation lags behind air temperature fluctuations at short time scales (e.g. daily basis), $T_{a_{smooth}}$ is the smoothed air temperature (^{o}C) represented as:

$$T_{a_{smooth}} = (1 - \lambda) T_a^{t-1} + \lambda T_a^t \quad (4)$$

where $T_a^{t-1}$is air temperature at time $t - 1$(^{o}C) and $T_a^t$is air temperature at time t (^{o}C).

Two different parameter sets for first-order headwater streams located either in the lowlands or in the highlands were established. For the lowland headwaters, $\mu = 0.5$, $\alpha = 20.5$, $\gamma = 0.2$, $\beta = 6.5$and $\lambda = 0.1$were chosen from the literature, considering the geographical similarities between the multiple basins analysed in Bawden et al. (2014).

In the Rocky Mountains, glacier melt dominates the Athabasca River flow regime and constitutes an important source of cold water into the system. Due to the lack of water temperature records at these locations, remote sensing tools were used to retrieve water surface temperatures from 13 cloud-free and snow-free locations. Landsat 5 and 8 thermal images were retrieved for the time frame between September 2003 and September 2016. Those images have a 30 m resolution, and are terrain-corrected level-1 products downloaded from USGS (2018). For Landsat 8 thermal images, only thermal band 10 (10.6–11.9 μm) was used, because thermal band 11 (11.50–12.51 μm) is significantly affected by stray light and thus not recommended for quantitative analysis (Montanaro et al., 2014).

The mono-window method developed by Qin et al. (2001) was used to derive water surface temperatures at Jasper Lake. The derived water temperature and the observed hourly air temperature at 10 am (Mountain time zone) at the Jasper Warden Weather Station (e.g., Fig. 1a) were used to fit the parameters in Equation (3) using the least squared error method (see Fig. 1b and c). The fitted model performed relatively well (Nash-Sutcliffe Efficiency (NSE) = 0.54, R^2 = 0.55) with low error (RMSE = 0.72 °C) and bias (PBIAS = 0.8%) after comparing simulated against retrieved surface water temperatures at Jasper

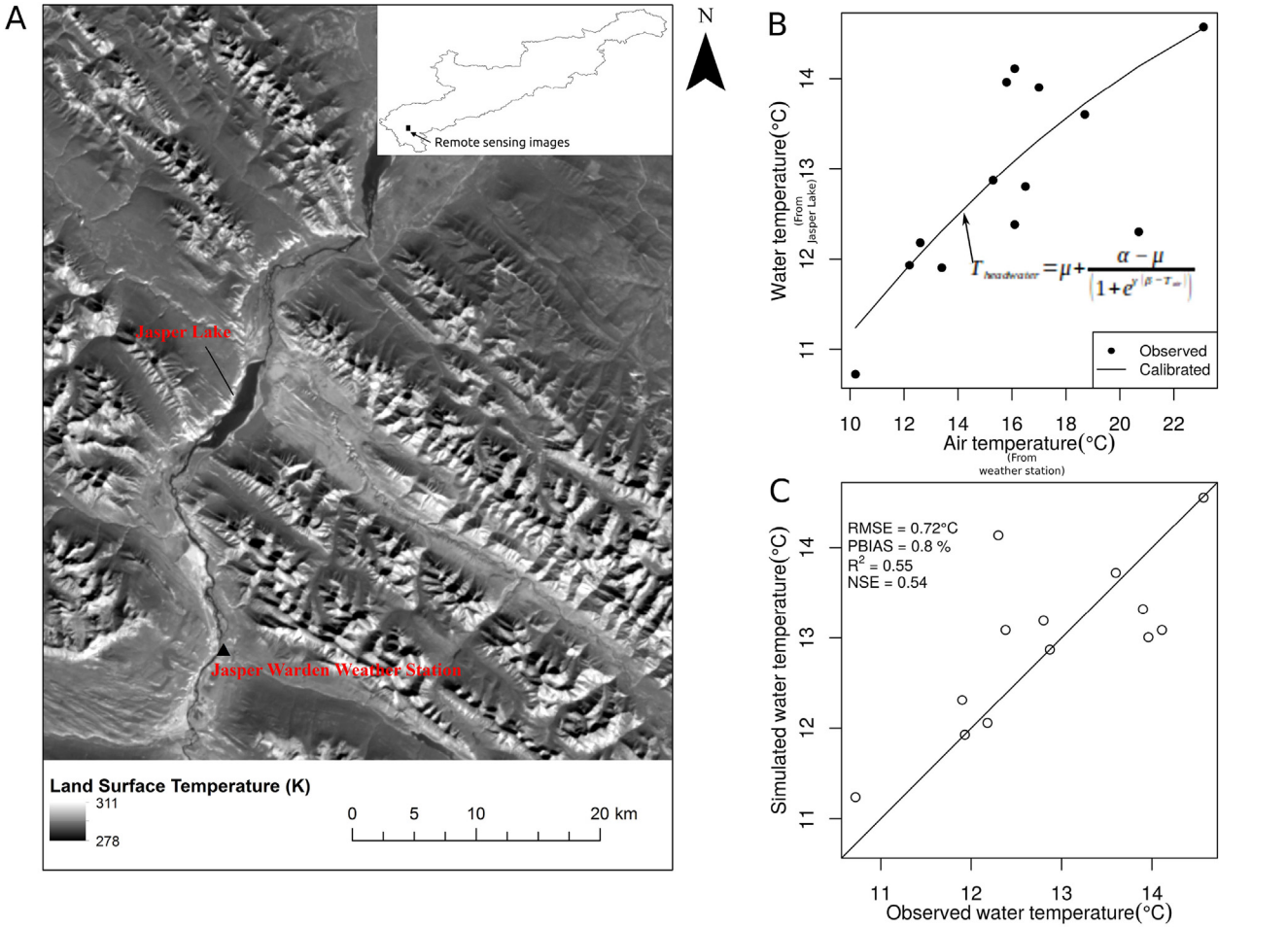

Fig. 1. a) Example of Landsat thermal image showing land surface temperature near Jasper Lake; b) observed air temperature from Jasper Warden weather station versus retrieved surface water temperature from Jasper Lake and Mohseni's water temperature regression model (see Equation (3)) and c) retrieved versus simulated surface water temperatures at Jasper Lake.

Lake (Fig. 1c). The resulting parameters $\mu = 0.5$, $\alpha = 16.28171$, $\gamma = 0.103247$, $\beta = 2.440703$and $\lambda = 0.1$were in agreement with the ranges established by Van Vliet et al. (2012). The calibrated parameters were used to set up the boundary conditions at all the headwater streams located upstream of the Hinton station.

2.3. Integration of RBM and MESH

RBM was designed to be used as stand-alone software or in conjunction with hydrological models such as VIC (e.g., Van Vliet et al., 2012) and DHSVM (Sun et al., 2015) to simulate nominal daily water temperatures in river basins. In this research, we coupled RBM with the MEC-Surface & Hydrology System (MESH) model (Pietroniro et al., 2007; Carrera et al., 2010) to first simulate hydrological processes and then stream temperature along the river reaches to extract the timing of freeze-up and breakup events in the ARB catchment (Fig. 2).

MESH is a semi-distributed physically-based land-surface hydrological modelling system developed by Environment & Climate Change Canada for hydrological applications (Pietroniro et al., 2007). It uses the Canadian Land Surface Scheme (Verseghy et al., 1993) for vertical exchanges and generation of lateral fluxes of energy and water for vegetation, soil and snow. WATROF is incorporated in the model for lateral movement of soil and surface water to the drainage system and the WATROUTE routes streamflow through the river channels using the Kinematic Wave approach (Kouwen, 1988). It also uses the Group Response Unit (GRU) approach, i.e. combining areas of similar hydrological behaviour, to reduce complexity and heterogeneity in the drainage basin for computational efficiency (Kouwen et al., 1993). This is a more suitable approach for large scale drainage basins due to its operational simplicity, while retaining the basic physics and behaviour of a distributed model (Pietroniro and Soulis, 2003). MESH has been successfully implemented in multiple large-scale cold region catchments that face river ice events every year (e.g., Davison et al., 2006; Haghnegahdar et al., 2014; Mekonnen et al., 2014; Davison et al., 2016; Yassin et al., 2017).

MESH requires seven meteorological input files: precipitation, air temperature, wind speed, barometric pressure, specific humidity, incoming longwave radiation and incoming shortwave radiation. Precipitation data was obtained from the Canadian Precipitation Analysis database, which is operationally run by the Canadian Meteorological Centre (CMC). The CMC produces accumulated precipitation for 6 h and 12 h time intervals over all of North America on a 10 km × 10 km grid (Fortin et al., 2015). Precipitation is interpolated using a geostatistical approach by combining CMC's short-range precipitation forecasts as the background field, a land-surface network of rain-gauge measurements, satellite observations and radar derived rain rates (Mahfouf et al., 2007). The rest of the gridded meteorological forcing data were obtained from the Global Environmental Multiscale Model (GEM), which is an integrated forecasting and data assimilation system developed by Environment Canada (Côté et al., 1998), available at spatial and temporal resolutions of 15 km and 1 h, respectively.

Our model coupling approach requires running both MESH and RBM models independently, with data from MESH being transferred to RBM. MESH and RBM were both set up on the same orthogonal grid

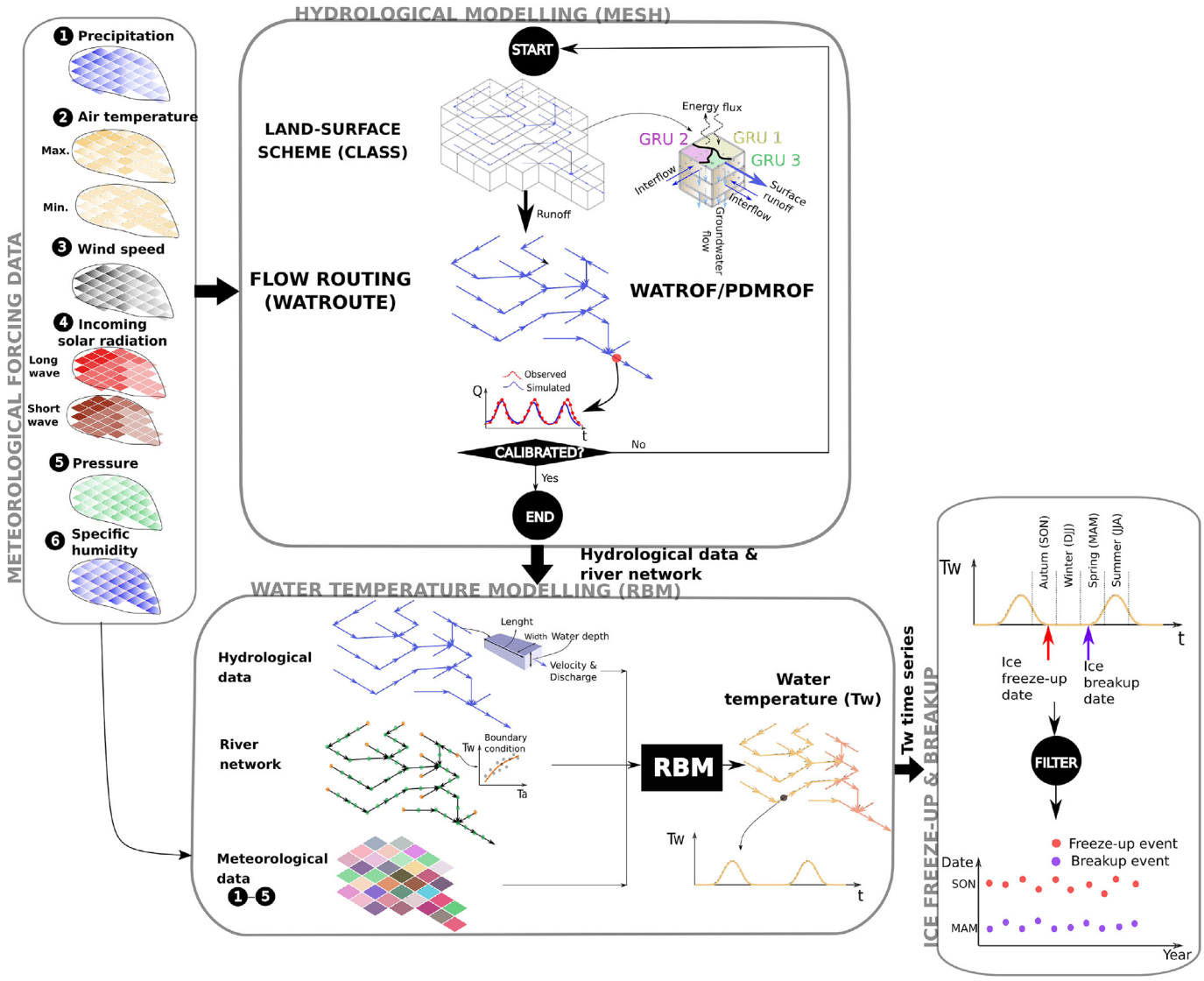

Fig. 2. Flow chart of the coupling of MESH and RBM models to simulate the timing of freeze-up and breakup events.

with a spatial resolution of 0.1250˚, resulting in 1326 grid cells and channels that comprehend the Athabasca River and tributaries. Firstly, MESH was run at a 3-h time-step for the period 01-Jan-2002 to 31-Dec-2012 to simulate streamflow, water depth, width and averaged water velocity in every channel. These hydrological and hydraulic variables, in a second step, were aggregated and introduced into RBM at daily time steps. Similarly to MESH, RBM requires meteorological variables at daily time steps. However, because meteorological observations are rarely discretized at the resolution of each segment/node, meteorological data were interpolated onto the river network. Additionally, river network topology data necessary for the computations were also introduced into RBM. Thirdly, RBM was run at daily time steps to compute water-depth averaged temperatures at every channel of the stream network. Finally, simulated water temperature time series were used to extract the freeze-up and breakup dates using a Gaussian linear filter algorithm to detect the inflection point of high water temperature gradients when the temperature drops to 0 ˚C in the fall and rises from the same temperature in the spring. The Gaussian filter uses a variable rectangular window that moves along the time series to smooth the noise associated to the rapid water temperature fluctuations aiding to detect the inflection points.

2.4. Study case

The Athabasca River basin is Alberta's largest unregulated river system, with a total drainage area of more than 150,000 km^2 (e.g., Cheng et al., 2017) (see Fig. 3). The basin extends across the Canadian prairie provinces of Alberta and Saskatchewan, starting at the Columbia Icefield in Jasper National Park (headwaters) and travelling 1538 km northeast across Alberta to drain into Lake Athabasca (e.g., Bawden et al., 2014).

The ARB has a mean annual temperature of 2 ˚C and average yearly precipitation of 460 mm, roughly 70% of which occurs as rain (e.g., Bawden et al., 2014; Canada, 2016). A snow-melt driven flow characterizes the hydrological regime of the basin, beginning in late April or May. Peak flows generally occur in June or July and are driven by snowpack runoff from the Rocky Mountains. The remaining months experience gradually declining flows until winter when low flow conditions resume in December (Burn et al., 2004).

Every year, the Athabasca River experiences multiple ice-jam formations and occasional flood events (e.g., Andres and Doyle, 1984; Peters et al., 2006). Ice-cover formation in the Athabasca River usually happens from October to November, while its breakup usually occurs at the end of March or beginning of April (Pietroniro et al., 1998; Hutchison and Hicks, 2007; She et al., 2009). Every year during the

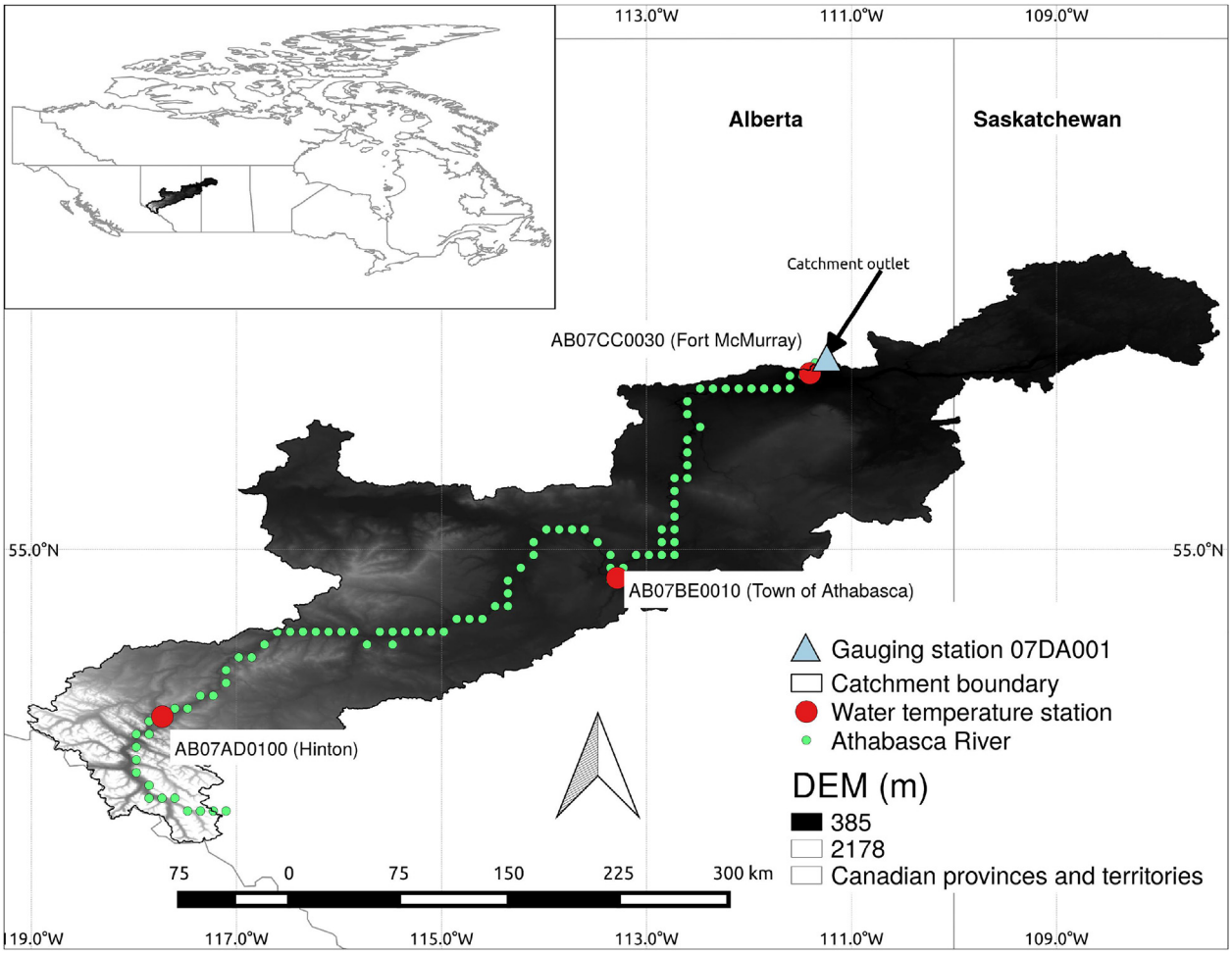

Fig. 3. Geographical location of the area of the Athabasca River basin upstream Fort McMurray. Streamflow and water temperature station locations, digital elevation model (DEM) and the Athabasca River model (green dots) are included. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

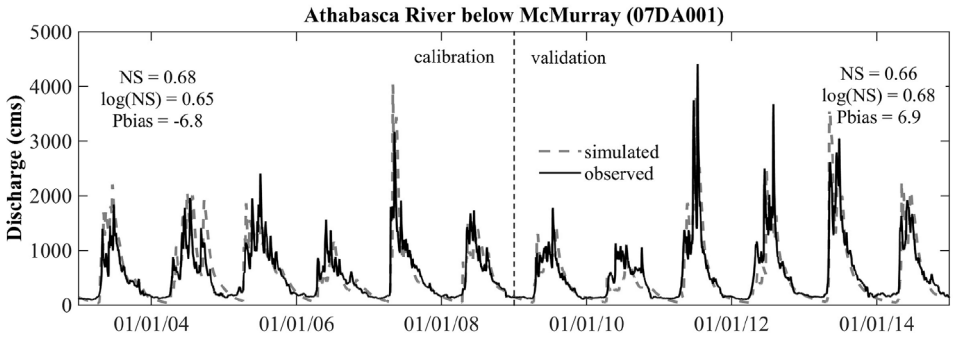

Fig. 4. Observed and simulated streamflows at the gauging station "Athabasca River below McMurray" (07DA001) for the calibration and validation period. MESH performance statistics are included.

breakup time, the risk of ice-jam formation and flooding is high in the Athabasca River. Severe floods in 1881, 1885, 1925, 1928, 1936, 1962, 1963, 1977 and 1997 have proven the importance of river-ice modelling.

3. Results

3.1. Calibration of the hydrology model

The calibration of MESH was performed using OSTRICH, which is an open source, auto-calibration and multi-algorithm parameter optimization software (Matott, 2005). Within OSTRICH, a parallel version of the Dynamically Dimensioned Search (PDDS) (Tolson and Shoemaker, 2008) algorithm was used to calibrate vegetation, soil, drainage and routing parameters from three dominant Group Response Units (GRUs). Though MESH was run at a 3-h time-step, output streamflows were generated at a daily scale for comparison and analysis. The model was first calibrated for the 2002–2008 period. The Nash-Sutcliffe Efficiency (NSE) was estimated to evaluate the model performance. Since NSE leads to an overestimation of the model performance during peak flows and an underestimation during low flow condition, the logarithm of the NSE (log(NS)) was estimated in order to

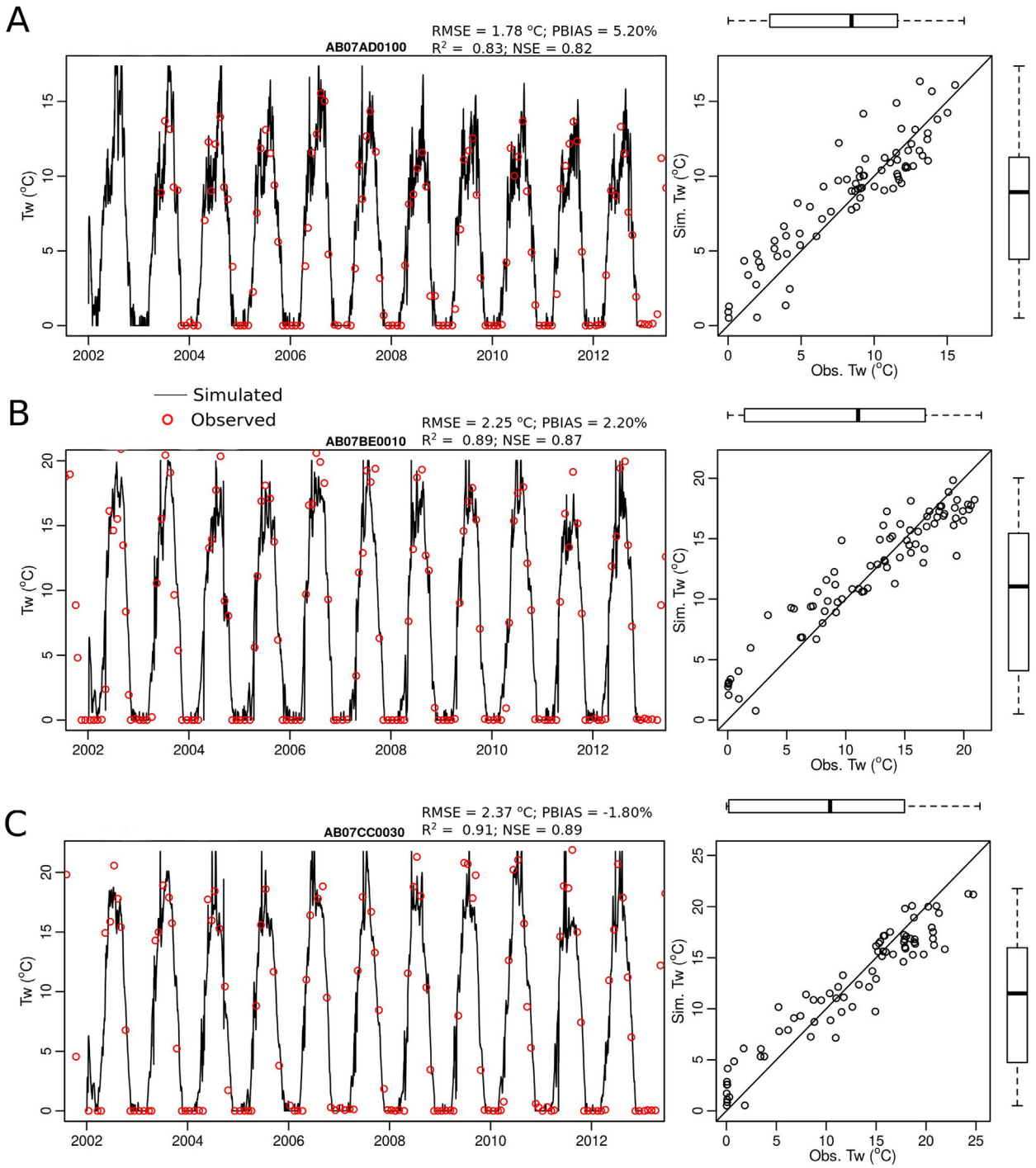

Fig. 5. Daily time series and scatter plots of observed and simulated stream water temperatures at the station a) AB07AD0100 (Hinton), b) AB07BE0100 (Town of Athabasca) and c) AB07CC0030 (Fort McMurray). RBM performance statistics are included.

evaluate the model integrally. The year 2002 was considered a spin up period and the remaining 6 years as the calibration period. Then the model was validated for 6 years, from 2009 to 2014. The percent bias (Pbias), which measures the average tendency of the simulated data to be larger (positive, overestimation) or smaller (negative, underestimation) than their observed counterparts, was also estimated.

Fig. 4 shows the results of the MESH calibration and validation of the flows recorded at the gauging station "Athabasca River below McMurray" (07DA001) (see Fig. 3 for station location), showing good agreement between simulated and observed flows. The NS value of 0.68, log(NS) value of 0.65 and Pbias of −6.8% were achieved for the calibration years, whereas for the validation years, a NS value of 0.66, log(NS) value of 0.68 and Pbias of 6.9% were obtained. Slight discrepancies between observed and simulated flows during the falling limb and the low flow period of the hydrographs are related to insufficient complexity of the baseflow algorithms in MESH.

3.2. Performance of stream temperature simulations

RBM was set up using the hydrological variables estimated in MESH and the meteorological forcing data mentioned previously. The model was run for the 2002–2012 period and 2002 was chosen as the spin-up

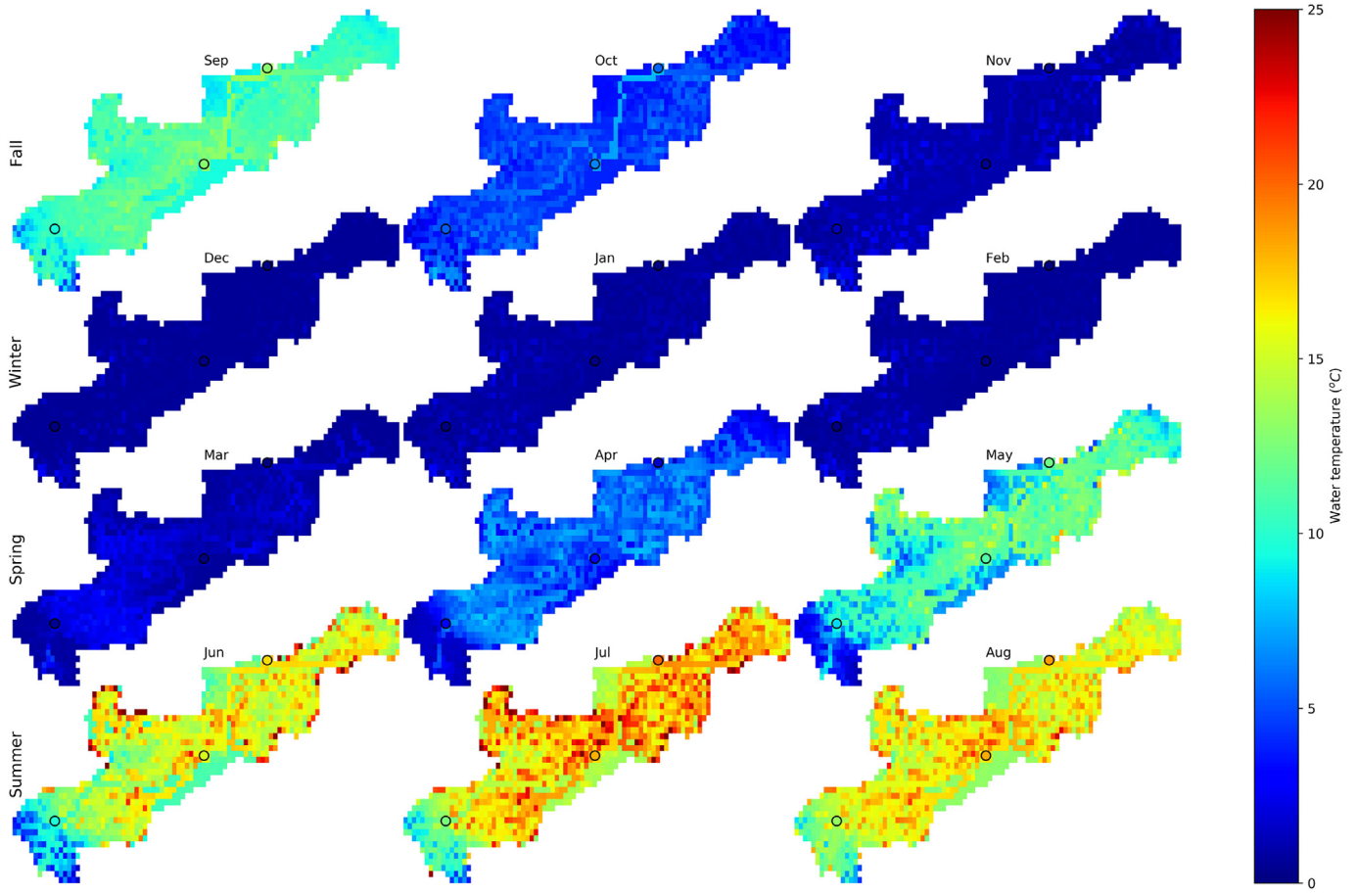

Fig. 6. Maps of multi-annual averages of monthly simulated water temperature (°C) in the ARB. The dots indicate observed water temperature at the station locations.

period. Daily water temperature simulations were compared to instantaneous water temperature observations unevenly obtained from Alberta Environment and Parks (http://aep.alberta.ca/water/reports-data/surface-water-quality-data/) at three different stations located on the Athabasca River (see Fig. 3 for station locations). These water temperature observations are instantaneous measurements taken approximately between 0 m and 1 m below the water surface around midday. The comparison between observed and simulated water temperature (Fig. 5) shows good agreement, with NSEs >0.8 and determination coefficients (R^2) >0.8 in all the stations revealing the model's ability to resemble the timing and magnitude of observed water temperatures. Although the model reproduces the daily water temperatures quite well, it tends to overestimate water temperature at the upstream and middle catchment sectors (Pbias >2.2%) and underestimate it (Pbias = −1.8%) at the outlet of the catchment. Water temperature overestimation in the highland tributaries can be attributed to the model's inability to fully capture the headwater cooling from snowmelt and glacier melt inflows. A characteristic of the simulated water temperature time series is the high frequency oscillation of the temperature along its rising limb during spring and at the peaks during summer. Such high frequency fluctuations represent model responses to rapid variations in the hydrological and meteorological forcing data.

3.3. Multi-annual stream temperature simulations

Stream water temperatures change spatially and temporally across the catchment. Warmer water temperatures are expected during the spring and summer months, when air temperatures rise over 25 °C in some tributary basins located at the mid-catchment area (see Fig. 6). In contrast, water temperatures decrease to zero degrees during late autumn, when the water surfaces freeze and remain frozen during winter and into early spring when the river ice breaks up. In April and May, water temperatures of tributaries in the Rocky Mountains are much lower and under 5 °C, due to continuous inflows of snow and glacier melt waters. However, during summer, water temperatures of the Rocky Mountains' tributaries rise continuously, reaching temperatures higher than 12 °C, until the end of September. After September, water cools rapidly in most of the tributaries except in the Athabasca River's main stem, where water temperatures are still around 10 °C due to large water discharges. Water temperatures decline to values close to 0°C in November and remain so during the winter.

At tributaries located in the mid-catchment, water temperatures are the highest, especially at small tributaries where streamflows are relatively low. When compared to air temperature fields, we observed that air temperature spatial patterns coincide with the simulated water temperatures, indicating the high correlation with meteorological forcing data. Moreover, this area is mainly covered by crops and grasslands that reduce stream shading and thus facilitate water heating. In such small tributaries, the energy balance at the water surface strongly controls the water temperature regime, indicating a larger dependency on meteorological forcing conditions than on streamflow (Yearsley, 2012). This leads us to conclude that boundary condition effects on water temperature rapidly decrease downstream and become very weak at the main river network tributaries, where thermal regimes are more influenced by the water mixing of inflowing tributaries.

Inter-annual maps of water temperature show some temporal changes across the catchments that reflect climate variability (see Fig. 7). During the 2002–2003 period, water temperatures appear to be

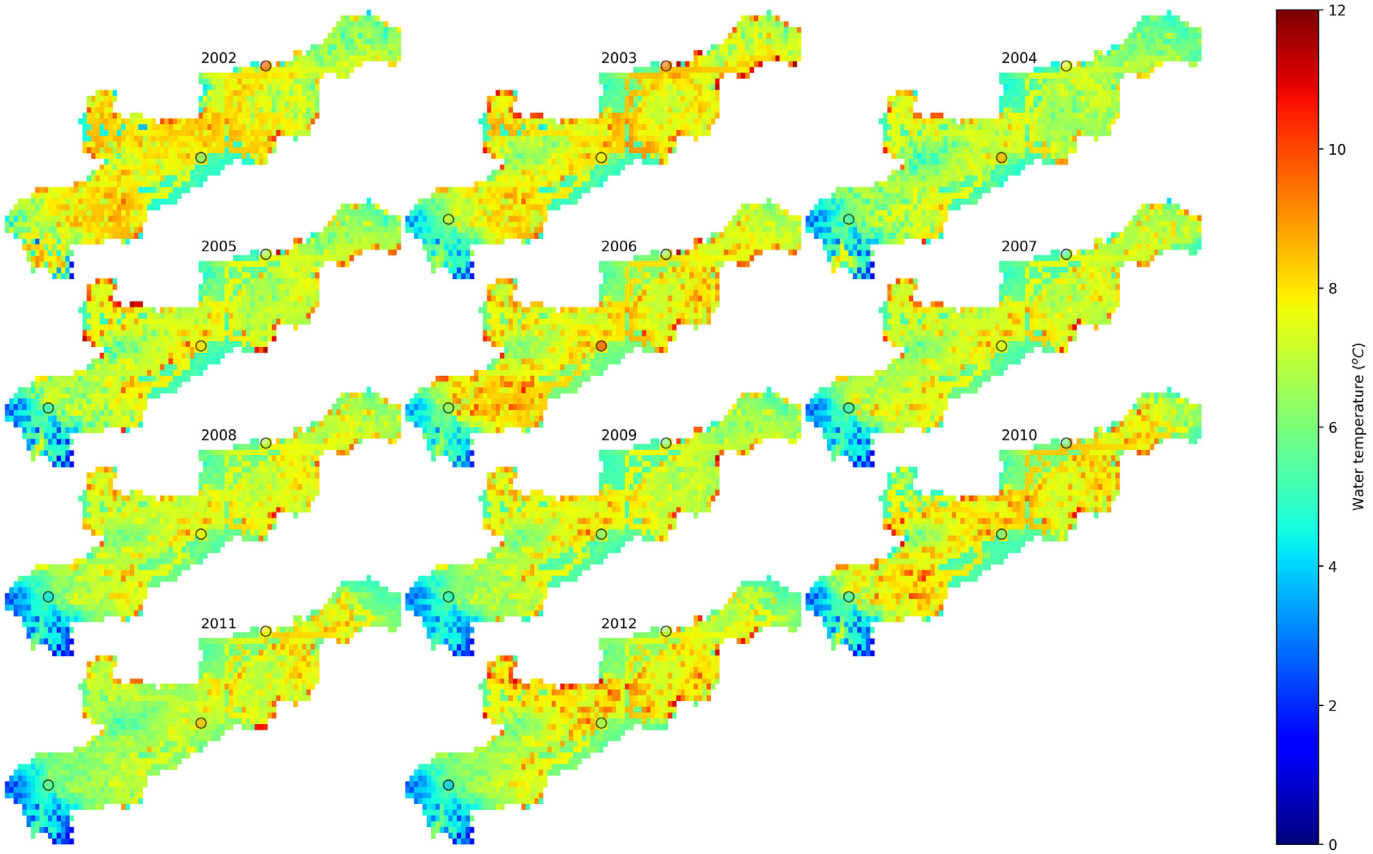

Fig. 7. Maps of multi-annual averages of simulated water temperature (°C) in the ARB. The dots indicate observed water temperature at the station locations.

the highest, particularly in the mid-catchment area, where annual averages were around 12 °C and coincided with intense droughts across the Canadian northern prairies (Bonsal and Regier, 2007; Yirdaw et al., 2008). During the subsequent years, water temperatures declined and were characterized by a relatively uniform pattern between 6 and 9 °C in the lowland tributaries and between 2 and 4 °C in tributaries in the Rocky Mountains. Similar patterns of uniform water temperatures were observed during 2007–2008 and 2011, which had been wetter years.

3.4. Validation of freeze-up and breakup timing

Freeze-up and breakup dates were extracted from the simulated stream water temperature time series across the catchment. Comparing the simulated freeze-up and breakup dates against observed B-flags (where B indicates backwater effect due to the presence of ice in the stream), recorded by Water Survey of Canada at the three analysed stations, we observed that the model was able to simulate these events with reasonable accuracy and an acceptable determination coefficient (R^2 range between 0.45 and 0.91) (see Fig. 8). As Fig. 8b and 8c shows, the model is able to simulate the breakup dates slightly better than the freeze-up dates with RMSE = 2 days and RMSE = 4 days, respectively. In general, the model performs better for the Town of Athabasca and Fort McMurray stations, where NSE are around 0.5 and 0.7 for the freeze-up and breakup events, respectively, than at Hinton station (NSE ≈0.41).

At the Hinton station, we observed that the model error for freeze-up and breakup dates was reflected in a RMSE of 5 and 10 days, respectively. Uncertainty is quite high in the simulated breakup dates, which is shown by a wider uncertainty band in the scatter plot. Difficulties in predicting breakup events in the Athabasca River were reported by Martin et al. (2013), who implemented a 2D hydrodynamic model and predicted early river-ice melt for some years. As compared to the other two downstream stations, earlier river freeze-up and ice-cover breakup occur at Hinton around the first weeks of November and April, respectively.

Temporal variability of the timing of freeze-up and breakup events is controlled by the interaction of climatic and hydrological factors. Increases in air temperature and relatively constant streamflow during 2006 and 2009 were responsible for delaying river water surface freeze-up at the towns of Athabasca and Fort McMurray by approximately 10 days. During the same period at Hinton, freeze-up timing did not change considerably, due to slight increases in streamflow that compensated for the air temperature rise. In contrast, air temperature reductions accompanied by streamflow reductions during 2005 and 2006 accelerated the freeze-up of the water surface within ~8 days at the three stations. Despite air temperature rises of around 2 °C, later freeze-up occurred during 2003 and 2004, particularly at the towns of Athabasca and Fort McMurray, due to significant increases in streamflow of more than 200%. During these two years, the timing of freeze-up was delayed by significant increases in streamflow and reductions in air temperature.

The timing of ice-cover breakup correlates ($\rho = 0.78$) with the air temperature changes. It is observed that when air temperature increases, early breakup occurs at all stations, while air temperature decreases delay these events. An acceleration of ice-cover breakup occurs from 2002 to 2004 at Hinton, from 2002 to 2005 at the Town of Athabasca and from 2002 to 2006 at Fort McMurray. A considerable spring air temperature rise of more than 3 °C was observed during these years at these stations. In contrast, streamflow fluctuations are slightly reduced at the Town of Athabasca and Fort McMurray. However, this is not the case at Hinton, where the rise in air temperatures clearly enhances the acceleration of breakup by around 15 days in 2002, even though there is a decline in streamflow from 80 to 40 $m^3 s^{-1}$. Similarly, Zachrisson (1989) reported the advancement of breakup events by

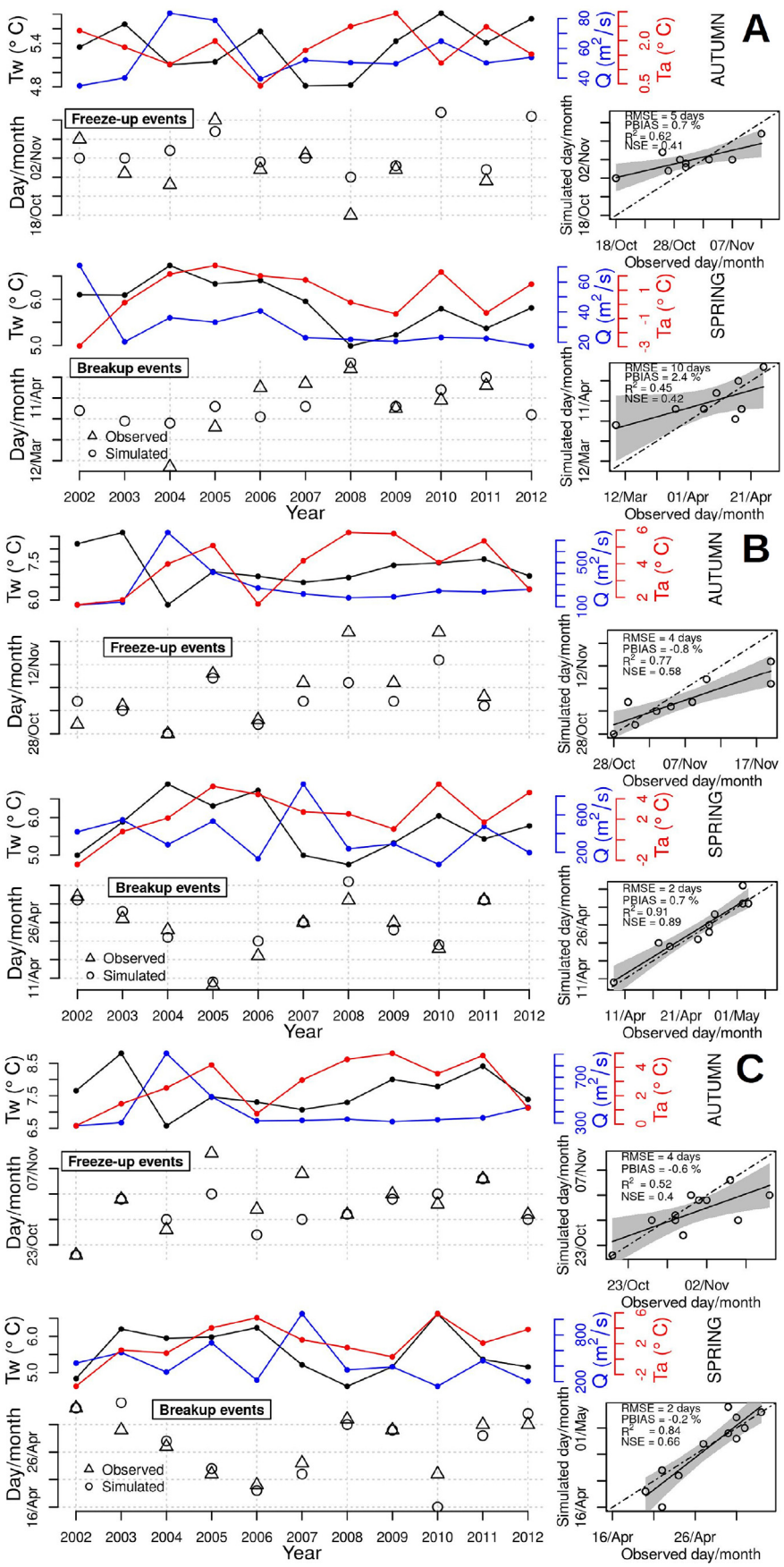

Fig. 8. Time series of freeeze-up and breakup dates, and scatter plots of observed and simulated freeze-up and breakup dates at a) AB07AD0100 (Hinton), b) AB07BE0100 (Town of Athabasca) and c) AB07CC0030 (Fort McMurray). Autumn and spring time series of averaged water temperature (Tw), air temperature (Ta) and streamflow (Q) are included for each station. Performance statistics and linear fitting curves are included in the scatter plots.

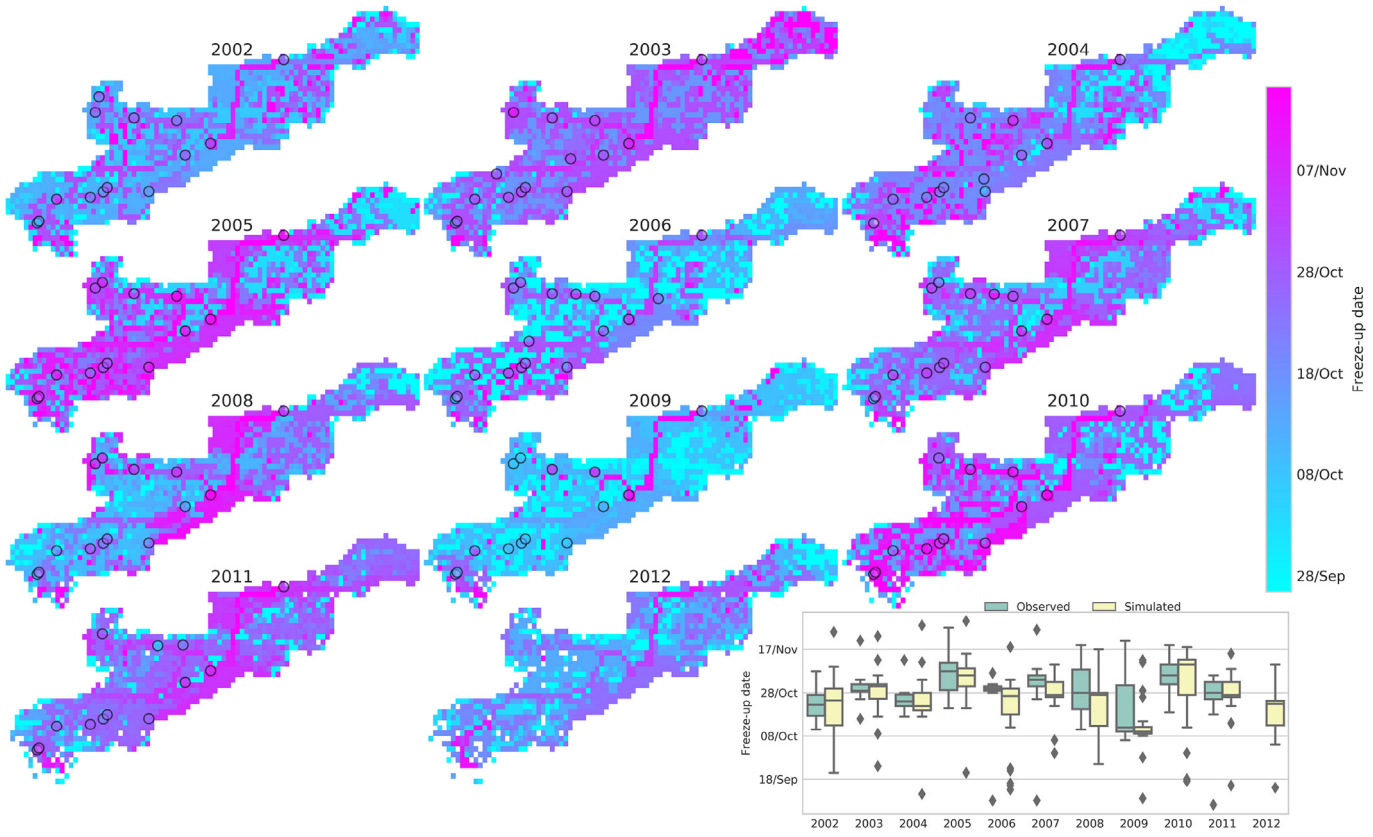

Fig. 9. Maps of freeze-up dates for the simulated years (2002–2012) and observed freeze-up dates at Water Survey Canada gauge stations (see overlaying dots). The inset shows boxplots based on observed and simulated freeze-up dates for each simulated year. Empty pixels in the maps represent non estimated freeze-up dates.

around 15 days after 3 °C air temperature increases in the Tornealven River in Finland, during the 1870–1950 period. In contrast, decreases in air temperature of more than 2 °C during 2005 and 2009, at the Town of Athabasca and Fort McMurray stations, delayed breakup occurrence by more than 10 days in 2005.

3.5. Spatio-temporal variation of freeze-up and breakup events

The timing of freeze-up and breakup events varies spatially across the catchment. Although freeze-up date maps (Fig. 9) do not show consistent spatial patterns, we can observe that freeze-up will always occur later at the main tributaries than in small streams influenced mainly by air temperature variability. Main tributaries froze over during early November and small tributaries in early October. However, in tributaries located in the mid-catchment, freeze-up occurs later due to higher air and water temperatures. At tributaries located in the Rocky Mountains in the north-western areas of the catchment, early freeze-up is expected due to lower air and water temperatures. For example, analysing the timing of freeze-up events along the Athabasca River (see Fig. 11), we observe, in general, that freeze-up advances in time with increases in air and water temperature and streamflow. At the headwater tributaries in the Rocky Mountains, freeze-up occurs during the second week of October. Freeze-up occurs later downstream and reaches a maximum freeze-up timing during the third week of November in reaches located in the foothills. Freeze-up timing occurs earlier, until the end of October, near Hinton but occurs progressively later in the flow direction quite smoothly and linearly until it reaches downstream of the Town of Athabasca, where freeze-up occurs consistently by the second week of November. Consistent freeze-up timing occurs until near the outlet, where decreasing air and water temperatures and increases in streamflow accelerate freeze-up during the first week of November.

The timing of breakup and freeze-up events vary, not only spatially but also temporally in the catchment. Comparing simulated and observed boxplots of freeze-up dates (see the inset in Fig. 9), we observed the capability of the model in capturing the temporal variability of freeze-up events throughout the analysed years. As mentioned before, variability in freeze-up timing is mostly associated with air temperature changes. Therefore, during 2002 and 2005, the freeze-up occurrence shifted from the third week of October to the first week of November. After 2005, freeze-up occurs earlier, with the earliest occurring around the second week of October in 2009. There is then a drastic delay in 2010, when freeze-up occurred in the first week of November. Freeze-up accelerates in the later years, occurring in the third week of October in 2012 due to a reduction in autumn air and water temperatures.

The spatial behaviour of breakup dates exhibits relatively clear patterns (see Fig. 10). At tributaries located in the Rocky Mountains, breakup events occur in late May; however, whereas downstream of Hinton, particularly in large tributaries, breakup occurs earlier, in late March and early April. Mid-May breakups are observed in tributaries located in north-eastern areas of the catchments and along main reaches of the Athabasca River. Such tributaries, where breakup occurs mostly in May, correspond to catchment areas with low air and water temperatures. Very early breakups are observed during mid-March, particularly in 2005, 2007 and 2009, at tributaries in the mid-catchment area between Hinton and the Town of Athabasca, due to high air and water temperatures. Analysing breakup timing along these reaches of the Athabasca River, we observed in general that breakup timing occurs earlier and is consistent with increases in air and water temperature and streamflow along the Athabasca River (see Fig. 11b). In the river, breakup begins along the foothill reaches near the Rocky Mountains in the first week of March. After 20 days, during the first week of April, breakup occurs nearly at the same time in reaches located between Hinton and the Town of Athabasca. Breakup advances

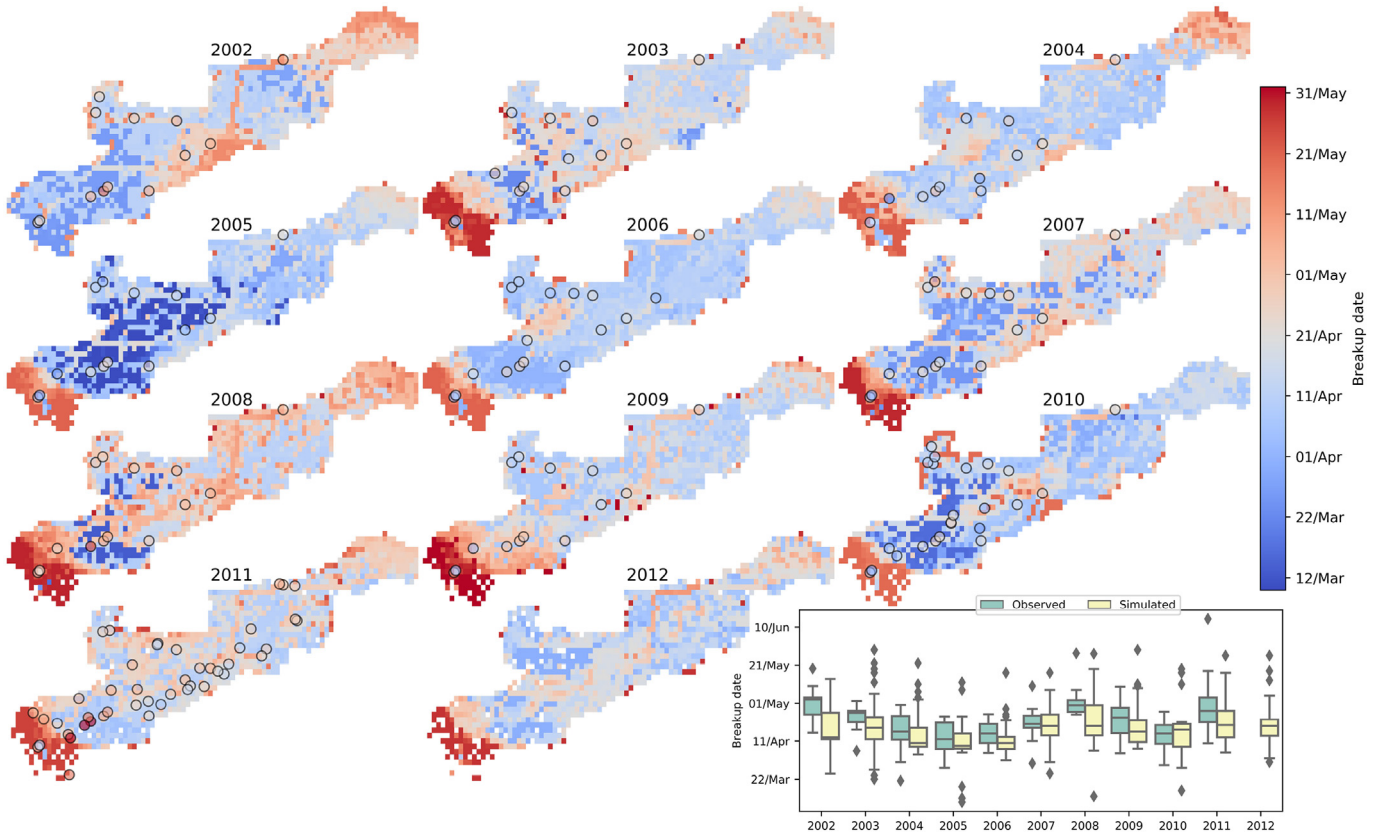

Fig. 10. Maps of breakup dates for the simulated years (2002–2012) and observed freeze-up dates at Water Survey Canada gauge stations (see overlaying dots). The inset shows boxplots plotted based on observed and simulated freeze-up dates for each simulated year. Empty pixels in the map represent non estimated breakup dates.

progressively until Fort McMurray, where breakup generally occurs during the second week of April. Late breakup occurs in mid June at the headwater reaches in the Rocky Mountains and just upstream of Hinton, where air and water temperatures are quite low and streamflows are close to $5\,m^3\,s^{-1}$.

Comparing box plots of observed and simulated breakup dates for the simulated years, we observed some discrepancies but similar variance in both data sets. As mentioned before, the model tends to underestimate the timing of breakup events, particularly in the lowlands. Despite the discrepancies, a periodic breakup timing behaviour controlled by air temperature variability is observed in both data sets. Breakup occurs progressively earlier from 2002 to 2006; later breakup events occurred until 2008. Earlier breakup occurred again from 2008 to 2010, remaining constant until 2012.

4. Discussion and conclusions

4.1. Estimation of the timing of river ice freeze-up and breakup events

The estimation of the timing of river ice freeze-up and breakup events involves many challenges. Computational models such as hydraulic models equipped with energy balance algorithms have been successfully utilized (e.g., Lindenschmidt et al., 2012a, b) to estimate stream water level, velocity and discharge necessary to trigger ice formation during autumn and ice breakup during spring at local scales. The validation of these models has always been challenging because the observed data sets are taken infrequently and rely on subjective opinions. Also, due to the scale of the model application, lack of data impedes the correct setup of model boundary conditions consisting of a streamflow hydrograph and a stream water temperature time series. Our hydrological and water temperature modelling approach represents a potential tool for estimating the timing of river ice freeze-up and breakup events on a large scale basis. In contrast to hydraulic river-ice models, this modelling approach shows the spatio-temporal variability, on a large scale, of how freeze-up and breakup events change. The model not only allows for the evaluation of the effects of climate and hydrological variability temporally, but also spatially, in order to understand those patterns that characterized the occurrence of these events. This approach presents a novel modelling system that could provide useful information to watershed managers regarding the timing of river freeze-up and ice-cover breakup within the catchment, in order to establish ice-jam flood mitigation plans.

In the past, different authors have analysed river ice records and spring and autumn climate data to understand the temporal and spatial variability of freeze-up and breakup events (e.g., Magnuson et al., 2000; Prowse and Bonsal, 2004). Commonly, the spatial and temporal variability of the air temperature 0°C isotherm has been studied regarding changes in freeze-up and breakup events and has been linked to the retreat or the advancement of the isotherm. For instance, Bonsal and Prowse (2003) found significantly earlier spring dates of the timing of the 0°C isotherm in western Canada, whereas the trend is weaker in eastern Canada. Prowse et al. (2007) reported a higher correlation between the spring 0°C isotherm and ice-off dates for rivers located in the interior cold regions of Canada, but lower correlation for warmer rivers located in maritime climate zones. Although these methodologies help us to understand the links between climatic drivers and river freeze-up and ice-cover breakup events, they do not consider the hydraulic and thermodynamic processes involved in these two events. Gray and Prowse (1993) indicated that south-to-north temperature gradients from the headwaters to the lowlands, are sometimes dominated by ice and glacier melt, and lowland streams in cold region catchments can impact breakup events substantially. Our modelling

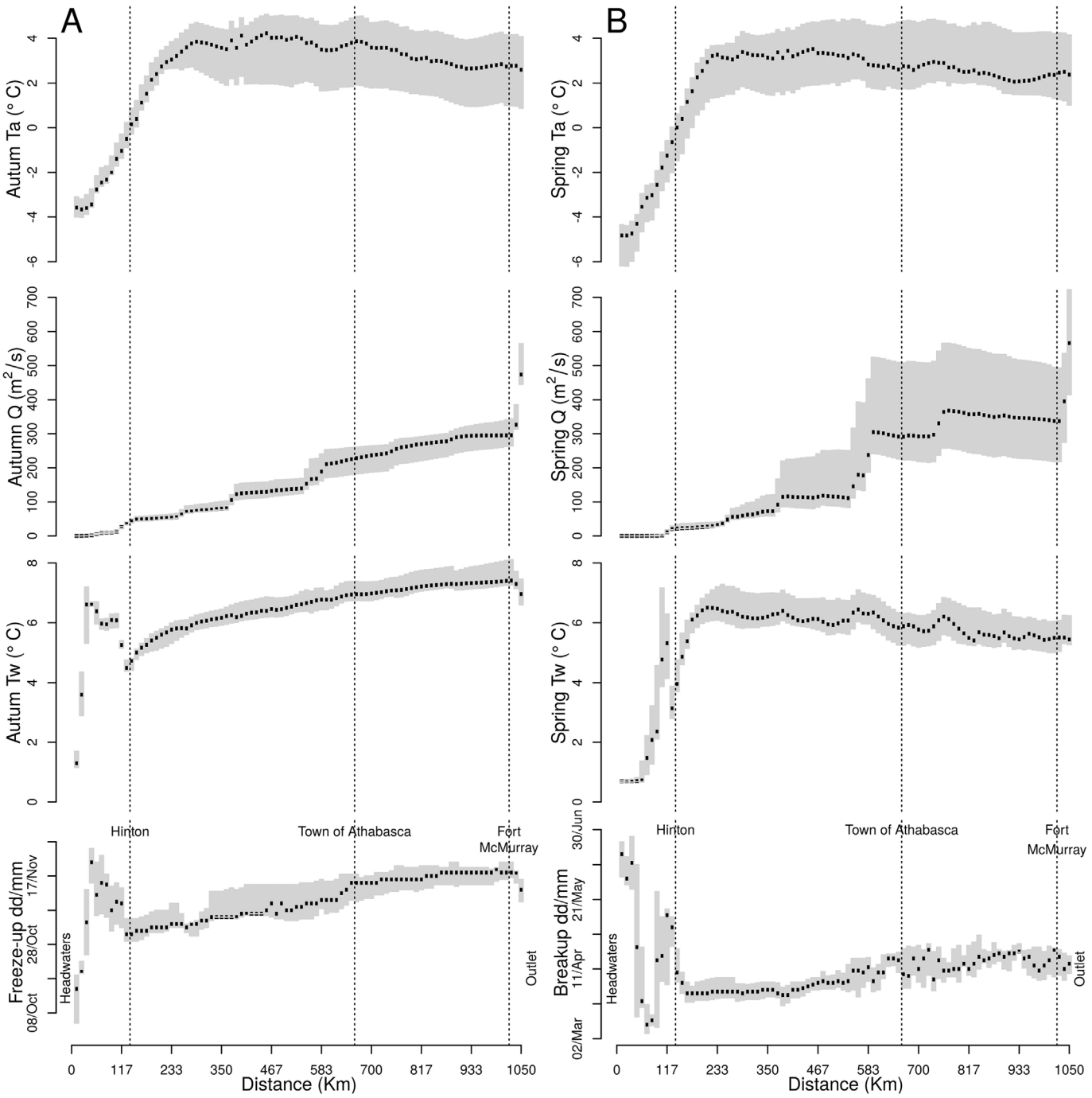

Fig. 11. Median and quartile range of annual averages of air temperature (Ta), streamflow (Q) and water temperature (Tw) along the Athabasca River model (see Fig. 3 for river location) for the median and quartile range of annual averages of a) freeze-up and b) breakup dates. Note that the 0 km distance indicates the river headwater reach whereas the most downstream river reach is at 1050 km.

approach combines meteorological, hydrological and hydrothermal drivers to estimate the timing of river freeze-up and ice breakup events, retrieved from stream water 0 °C isotherm simulated after coupling a physically-based hydrological model and a water temperature model.

In general, our model shows that, in the main tributaries of the ARB, freeze-up timing spans from the last week of September to the second week of November and becomes progressively later non-linearly from the headwater reaches to the catchment outlet. The breakup timing spans from the second week of March to the last week of May and occurs first at headwater reaches and lastly at the intermediate and outlet reaches. The model reports high correlation between freeze-up and breakup dates with meteorological and hydrological variables, also showing that such events are spatially controlled by non linear increases in air and water temperatures and streamflows along the streams. Although it is known that the formation of river ice is driven by atmospheric and hydrological conditions (Prowse and Beltaos, 2002; Prowse, 2006), we do not know in which proportion these two factors affect river freeze-up and ice-cover breakup. Our model shows that, for instance, spatial negative (positive) gradients of air temperature can trigger later (earlier) freeze-up events in contiguous catchment tributaries. Particularly in the Athabasca River, slightly negative gradients of air temperature near the catchment outlet downstream of Fort McMurray can cause earlier freeze-up despite relatively substantial streamflow increases. According to the model results, it seems that

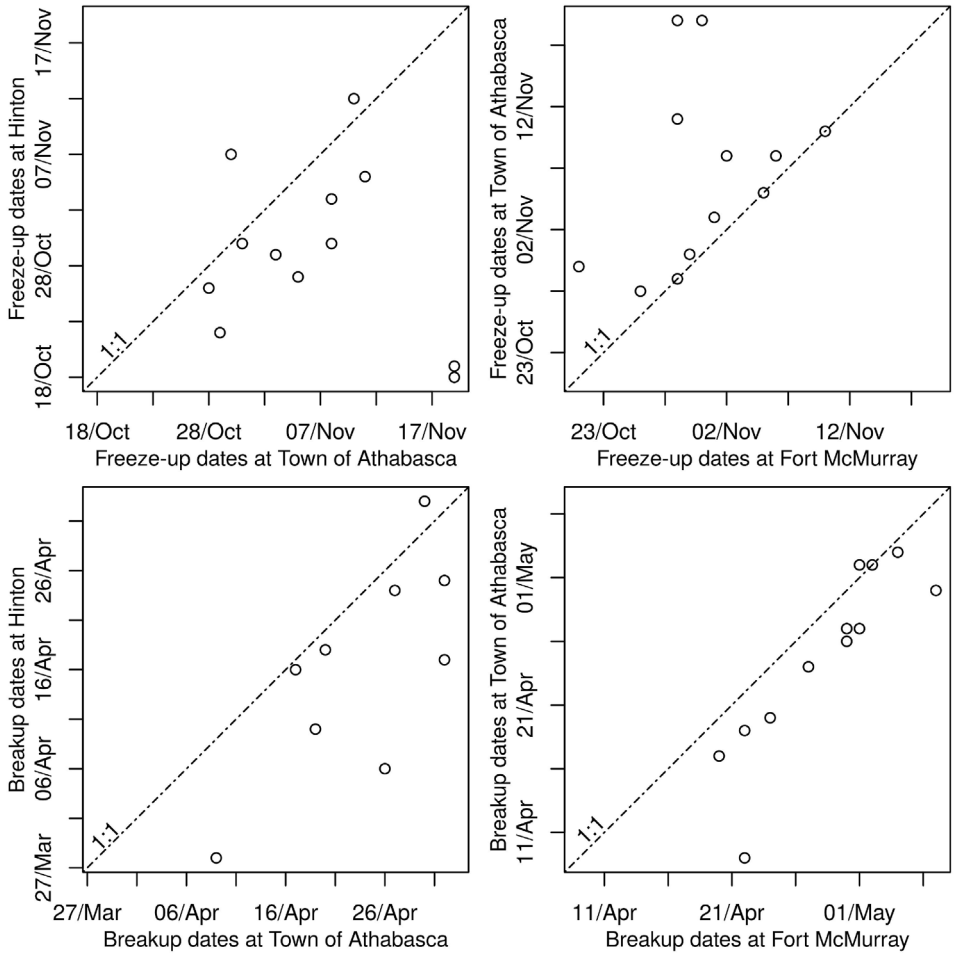

Fig. 12. River freeze-up dates (upper panels) and ice-cover breakup dates (lower panels) for the years between 2002 and 2013, comparing Hinton with the Town of Athabasca (left panels) and the Town of Athabasca with Fort McMurray (right panels).

freeze-up timing correlates better and is more sensitive to changes in air temperature than to changes in streamflow in the catchment. Results reported by Ginzburg (1992) and Soldatova (1992) indicate strong correlations between autumn and spring air temperature and freeze-up and breakup timing, respectively.

Although there is a high correspondence between observed and simulated breakup timing at the stations in the towns of Athabasca and Fort McMurray, breakup events show larger uncertainty in other regions within the catchment. Smith (2000) reported a lack of significant trends in breakup events that might be related to inter-decadal variability and inconsistent data length (Prowse and Bonsal, 2004). Other sources of error may be related to the fact that different criteria have been applied to recorded breakup events. For instance, some agencies report the dates of breakup initiation whereas others record the date of ice clearance at the end of breakup. Since differences between these two dates can sometimes be as much as 4 weeks, that may represent a large source of discrepancies.

4.2. Spatial variability of river ice freeze-up and breakup events

Theoretically, at freeze-up, an ice cover will initiate at a bridging, for example at the basin outlet, and extend towards the headwaters by juxtapositioning, when frazil ice that is generated in the upstream open-water stretches accumulates at the ice-cover front to extend it in the upstream direction. However, it is quite common that bridgings occur at several points along the river, often at river width constrictions, sharp meanders, bridge piers or islands, to form many ice covers simultaneously during the river freezing period. This was observed along the Dauphin River in Manitoba when four ice covers were formed along a 40 km stretch in just a few days during a cold bout (Lindenschmidt et al., 2013). Hence, it should not be a surprise to observe freeze-up dates often occurring earlier at Hinton than at the Town of Athabasca (see Fig. 12, top left panel) with most of the freeze-up date points lying below the 1:1 line, indicating an earlier freeze over of the Athabasca River at the Hinton reach than the reach at the Town of Athabasca. Many of the freeze-up date points for the Town of Athabasca and Fort McMurray lie on the 1:1 line (Fig. 12, top right panel), indicating that the freeze over of the river stretch between these two towns can be quite rapid, which is also reflected in the profiles of Fig. 11. However, freeze over can be later at the Town of Athabasca than at Fort McMurray, which is captured in the uncertainty band of the profile in Fig. 11. Better accuracy of the freeze-up dates may be obtained when a river ice hydraulics model, for example RIVICE (Lindenschmidt, 2017), is coupled to the MESH-RBM system, a topic of future work. For initial steps of this research please see Das et al. (2017), Rokaya et al. (2017) and Rokaya et al. (accepted).

For rivers flowing in a northerly direction, breakup theoretically begins at the headwaters and works its way downstream towards the basin outlet. Breakup consistently occurred earlier at the Town of Athabasca than at Fort McMurray (see Fig. 12, bottom right panel), on average approximately 3–4 days earlier. This is reflected in the breakup profile in Fig. 12. However, the breakup dates between Hinton and the Town of Athabasca are quite variable, sometimes Hinton experiences an ice-cover breakup earlier, sometimes later than at the Town of Athabasca. The variability in the breakup times between the two towns can be quite substantial, up to several weeks difference, which is also

captured in the uncertainty band of the breakup dates profile in Fig. 11.

4.3. Model limitations and future work

Some model limitations need to be addressed for future work. Although the proposed framework provides a flexible approach to feeding MESH output results into the RBM stream water temperature model, there are some key sources of uncertainty that will need to be kept in perspective while applying the methodology. The four primary sources of uncertainty in the current study are: 1) water temperature boundary conditions at headwater reaches, usually dominated by snow and glacier melt waters; 2) lack of consideration of anthropogenic impacts along the main catchment streams; 3) propagation of errors from MESH to RBM and 4) meteorological forcing data uncertainties, particularly, those associated with precipitation products. We believe that these uncertainties introduce errors in the estimation of freeze-up and breakup events, which need to be considered in future work. To address the first point, we implemented remote sensing techniques to extract water temperatures in mountainous areas to impose boundary conditions at the headwaters. Although acceptable results were obtained, more effort needs to be made in order to obtain longer records and apply them in areas which lack water temperature records. In relation to anthropogenic impacts in catchments, MESH has been substantially improved in order to consider streamflow regulation and water abstraction, factors that modify hydrological functioning in downstream waters, affecting river ice dynamics. Errors in streamflow and hydraulic variables introduced in RBM and estimated in MESH need to be reduced by a better discretization of both the catchment domain and meteorological forcing data.

Future modelling work using the approach described here will investigate climate change scenarios to predict ice phenology. Estimation of RBM boundary conditions using the CLASS energy balance computations is worthy to be tested in future works in order to facilitate the imposition of boundary conditions in headwater catchments. This modelling approach is potentially useful for decision support (for example, for water quality planning, hydropower management and resource management on a large scale), risk analyses of ice-jam floods and to study climate change impacts of large river basins around the world.

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